A Stratiform Cloud Parameterization for General Circulation Models

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The crude treatment of clouds in general circulation models (GCMs) is widely recognized as a major limitation in applying these models to predictions of global climate change. The purpose of this project is to develop a parameterization for stratiform clouds in GCMs that expresses stratiform clouds in terms of bulk microphysical properties and their subgrid variability.

In this parameterization, precipitating cloud species are distinguished from non-precipitating species, and the liquid phase is distinguished from the ice phase. The size of the non-precipitating cloud particles (which influences the cloud radiative properties and the conversion of non-precipitating cloud species to precipitating species) is determined by predicting both the mass and number concentrations of each species.

Cloud Microphysics

The stratiform cloud parameterization is based on a bulk cloud microphysics parameterization originally developed at Colorado State University for mesoscale cloud models (Tripoli and Cotton 1980; Cotton et al. 1982; Cotton et al. 1986; Meyers et al. 1992). We first improved the computational efficiency of the parameterization by introducing two approximations appropriate for stratiform clouds (Ghan and Easter 1992), so that the parameterization can now be applied to GCMs.

To permit application to both polluted continental clouds and pristine marine clouds, we have introduced the droplet number concentration N_c as a prognostic variable:

$$\frac{\partial N_c}{\partial t} = -\nabla\left(N_c \mathbf{V}\right) + N_{vc} - E_{cr} - A_{cr} - C_{cr} - C_{ci} - C_{cs} - F_{ci} \qquad (1)$$

Here $\mathbf{V}$ = three-dimensional velocity

N_{vc} = the rate of droplet nucleation

E_{cv} = the droplet evaporation rate

A_{cr} = the rate of autoconversion of cloud droplets to rain

C_{cr} = the rate of collection of cloud droplets by rain

C_{ci} = the rate of collection of cloud droplets by ice

C_{cs} = the rate of collection of cloud droplets by snow

F_{ci} = the rate of freezing of supercooled cloud droplets to form cloud ice.

Most of the sink terms in the droplet number balance follow from the sink terms in the cloud water mass concentration, assuming the sink processes affect the cloud water mass and number concentration, but not the average droplet mass. The droplet number sink due to autoconversion of cloud water to rain is parameterized according to Ziegler (1985). We assume evaporation depletes droplet number only when cloud water in a layer decreases to zero.

The droplet source reflects the nucleation of cloud droplets when aerosols are activated as cloud condensation nuclei. If we assume droplets are formed only as air enters a cloud, then the droplet source can be expressed

$$N_{vc} = -\nabla \cdot (N\mathbf{V}) \tag{2}$$

where N is zero except for inflow on the cloud boundaries, when it equals the number concentration of aerosols activated. The droplet source term represents a flux convergence of droplets into the cloud, which is not accounted for by the transport term $\nabla \bullet (N_c \mathbf{V})$ because the treatment of transport assumes no droplets flow into the cloud.

In applying (2) to the prognostic equation for droplet number, we have found that turbulent variations in velocity V must be considered if the prognostic equation for droplet number also includes a treatment of turbulent transport that, like the treatment of resolved transport, assumes no droplets flow into the cloud. Thus, if the turbulent transport is expressed in terms of a vertical diffusivity K, the droplet nucleation term for a layer at the base of a Euclidian (rectangular) stratiform cloud can be written

$$N_{vc} = \frac{\left[max(w_b, 0) + \frac{K_b}{\Delta z} \right] N_b}{\Delta z} \tag{3}$$

where w is the vertical velocity, Δz is the model layer thickness, and the subscript b denotes cloud base. Note that we neglect droplet nucleation on the sides of the clouds.

To determine the number concentration of droplets nucleated, N_n, we have developed a parameterization in terms of the vertical velocity and the aerosol number concentration, N_a:

$$N_n = \frac{w^* N_a}{w^* + cN_a} \tag{4}$$

where c is a coefficient that depends on the temperature, pressure, aerosol composition, and the mode radius and standard deviation of the aerosol size distribution (Ghan et al. 1993), and

$$w^* \equiv w + \frac{\alpha}{\alpha + \beta} \left[\frac{c_p}{g} \frac{T}{\theta} \frac{1}{\rho} \nabla \cdot \overline{\rho \theta' V'} - \frac{Q_r}{g} - \frac{1}{\alpha} \frac{c_p}{g} \frac{1}{\rho} \nabla \cdot \overline{\rho q_v' V'} \right] \tag{5}$$

with $\alpha \equiv \frac{\partial q_s}{\partial T}$, $\beta \equiv \rho c_\rho \frac{\partial q_s}{\partial p}$ and Q_r the radiative heating rate (Ghan et al. 1993).

Note that with the general expression (5), the parameterization accounts for supersaturation forcing by radiative cooling, turbulence moistening, and turbulent cooling. We have compared the number nucleated according to (4) with that simulated by a detailed size-resolving nucleation model (Edwards and Penner 1988) and have found that, even for realistic aerosol size distributions, the number nucleated agrees to within 30% for vertical velocities ranging from 1 to 500 cm s^{-1} and aerosol number concentrations ranging from 50 to 5000 cm^{-3}.

The parameterization (4) is restricted to the case of a single aerosol type. We have now developed a parameterization applicable to the more general case of activation of multiple aerosol types, with different compositions or size distributions. The different aerosol types can compete with each other as cloud condensation nuclei. Figure 1 compares the parameterized and simulated number fraction of aerosols activated for each of two competing aerosol types. The number concentrations and size distributions of both aerosol types are identical, but type 1 aerosol is fully soluble while type 2 aerosol is composed of 10% soluble material. The parameterization correctly predicts the more efficient activation of the more soluble aerosol type, with errors in the fraction activated between 20% and 50%. Additional comparisons are reported in Ghan et al. (**in press**).

Subgrid Cloud Parameterization

Subgrid scale variations in cloud microphysical processes must be accounted for in GCMs because cloud processes are highly nonlinear and are poorly resolved by the coarse grid size of GCMs. We express subgrid variations in cloud properties in terms of idealized probability distributions of the cloud variables. We assume most subgrid variability in stratiform clouds is due to turbulence and use the

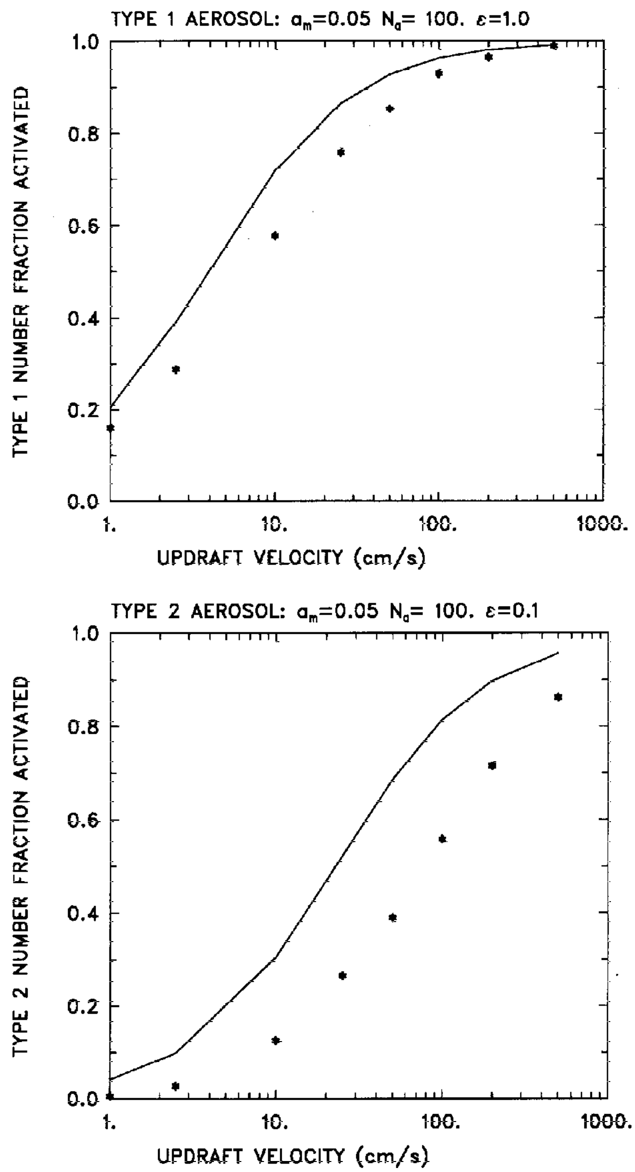

Figure 1. Number fraction of aerosols activated for two competing aerosol types, as simulated (asterisks) and as parameterized (solid line), plotted as functions of the updraft velocity. One aerosol type (top) is fully soluble, while the fraction of soluble material of the other (bottom) is 10%.

Mellor-Yamada second-order turbulence closure scheme to predict the variance of cloud variables. We account for subgrid cloud variations in cloud processes by integrating the expressions for the cloud processes over the probability distributions of the cloud variables. For example, the flux of cloud droplets at cloud base is expressed

$$\overline{wN'} = \int_0^\infty w\, N\, P(w)\, dw \tag{6}$$

where $P(w)$ is the probability distribution of vertical velocity, determined from the predicted mean and variance of vertical velocity.

Application to a Single-Column Model

To test the subgrid cloud parameterization, we have applied it to a single- column model. In addition to cloud microphysics and turbulence, the model treats vertical advection, radiative transfer, and surface processes. Figure 2 illustrates a simulation of a boundary layer cloud driven only by radiative

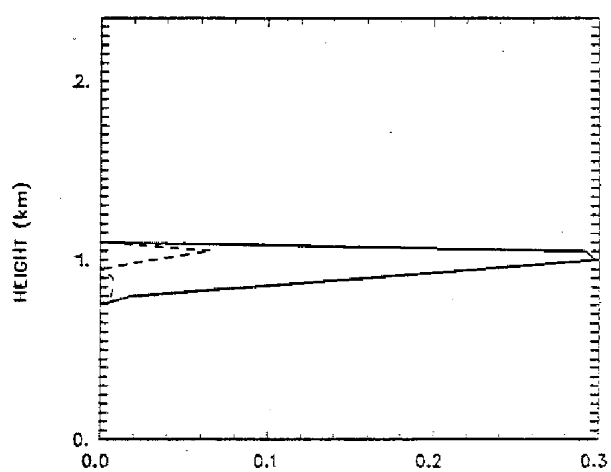

Figure 2. Mean cloud water (solid line) and standard deviation of cloud water (dashed line) as functions of altitude for a stratocumulus cloud driven by radiative cooling.

cooling. The subgrid standard deviation in cloud water is greatest near cloud top, where strong radiative cooling drives turbulent mixing but is much smaller than the mean cloud water at all levels. Consequently, the cloud fraction is unity throughout the depth of the cloud. With droplet number concentration prescribed at 300 cm^{-3}, no precipitation forms because subgrid variability in autoconversion is not yet treated.

Application to a GCM

We have applied the bulk cloud microphysics parameterization to the Pacific Northwest Laboratory (PNL) version of the National Center for Atmospheric Research (NCAR) Community Cloud Model (CCM1). We have replaced the usual prognostic variables temperature T and water vapor mixing ratio r_v with the condensation-conserved variables $T_{cld} = T - L/c_p r_c$ and $r_w = r_v + r_c$, where L is the latent heat of condensation. Temperature, water vapor, and the cloud water mixing ratio r_c can be diagnosed from T_{cld} and r_w by assuming condensation instantaneously eliminates supersaturations with respect to liquid water. Advection of cloud water is implicitly treated in the advection of T_{cld} and r_w and therefore need not be treated explicitly, thus eliminating problems associated with advecting a field with frequent zeroes.

Subgrid variations in stratiform clouds are not yet treated, but detrainment of condensed water from cumulus clouds is. We have performed one preliminary 30-day simulation with prescribed droplet number and have found that, without any tuning, the simulated global planetary radiation balance is within 10 W m^{-2} of satellite observations for both solar and infrared radiation (Table 1). Figure 3 shows the latitudinal distributions of zonal mean simulated and observed planetary radiation balance. Simulated cloud cover is high, about 80%, because the model simulates extensive ice clouds that are too thin to be observed.

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Table 1. Planetary Radiation Balance (July).

Radiation	Simulated W m^{-2}	Observed W m^{-2}
Outgoing Longwave		
clear-sky	269	268
total	228	238
cloud forcing	41	30
Absorbed Solar		
clear-sky	284	281
total	240	234
cloud forcing	-45	-46
Net Earth Radiation Budget		
clear-sky	15.3	13.5
total	11.5	-3.1
cloud forcing	-3.8	-16.6

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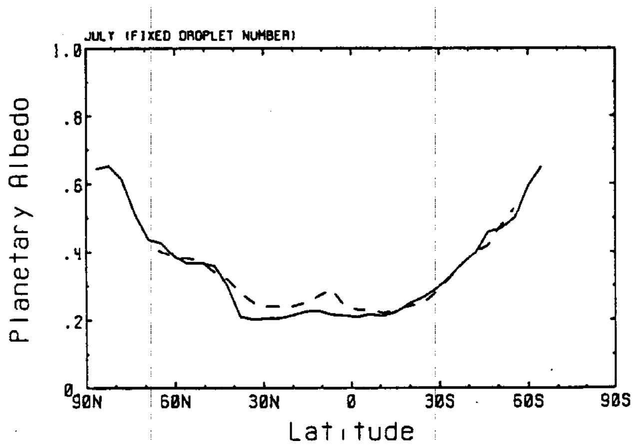

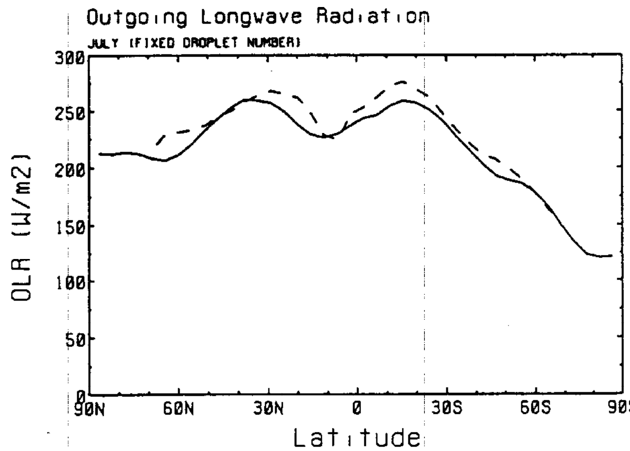

Figure 3. Zonal mean planetary albedo (left) and outgoing longwave radiation (right) as simulated by the PNL version of the NCAR CCM1 (solid line) and as observed by the Earth Radiation Budget Satellite (dashed line) for July.

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